On the Relationship between Isostatic Elevation and the Wavelengths of Tectonic Surface Features on Venus

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Regional isostatic elevation ($d$) and the widths and spacings (wavelengths, $\lambda$) of tectonic features formed due to horizontal extension and compression are observed quantities on Venus that are sensitive to the thermal and compositional structure of the lithosphere. To investigate the relationships between these parameters and the implications for Venus' lithosphere structure, we formulate models in which $d$ and $\lambda$ are functions of surface thermal gradient ($dT/dz_0$) and crustal thickness ($c$). In models that relate $\lambda$ to $dT/dz_0$ and $c$, we impose a requirement that for tectonic features with multiple wavelengths of deformation to develop, the growth rates of instabilities that control the dominant wavelengths must be "similar" in magnitude. This constrains the maximum values of $c$ and $dT/dz_0$ that can result in multiple tectonic wavelengths on Venus to be less than or equal to those determined in our previous study, $c \leq 30$ km and $dT/dz_0 \leq 25$ K km$^{-1}$ (M. T. Zuber, 1987, J. Geophys. Res. 92, E541-E551). We also estimate the strength of Venus' upper crust in areas with multiple wavelengths and obtain limits that are most consistent with a compressional origin for ridge belts, which are enigmatic features that have been purported to be of both extensional and compressional origin. In models that relate $d$, $\lambda$, $dT/dz_0$, and $c$, we show that in areas of Venus where the upper mantle is stronger than the upper crust, the spacings of short-wavelength ($\lambda \approx 30$ km) features should increase with increasing $d$ if the change in $d$ is due to increasing $c$, but should decrease with increasing $d$ if the change in $d$ is due to increasing $dT/dz_0$. Long-wavelength ($\lambda \geq 30$ km) features are not strongly correlated to variations in $d$ due to the effect of crustal thickness on the depth distribution of lithospheric strength, which influences wavelength selection in a complex manner. In areas of Venus where the upper mantle is weaker than the upper crust, the spacings of short-wavelength features should remain constant with increasing $d$ if the change in $d$ is due to a change in $c$, but should decrease with increasing $d$ if the change in $d$ is due to increasing $dT/dz_0$. The results demonstrate that in shallowly compensated areas of Venus that contain regularly spaced tectonic features, it may be possible to distinguish whether isostatic elevation differences are a consequence of variations in crustal thickness or thermal gradient. We interpret the observed relationship of $\lambda$ with $d$ for short-wavelength tectonic features in several tessera regions to reflect crustal thickness variations.

INTRODUCTION

One of the major outstanding questions concerning the evolution of Venus or any other planet is the manner in which heat is transported from its interior. Current interpretations for Venus range from suggestions that the primary mechanism of heat transport is by hot spots (Morgan and Phillips 1983, Grimm and Solomon 1987, Kiefer...
and Hager 1988) or plate recycling (Head and Crumpler 1987) or is unknown (Solomon and Head 1982). One basis for distinguishing between these hypotheses is an understanding of the thermal and mechanical structure of the lithosphere. Given the available data for Venus, one means of obtaining direct information about the lithosphere is from analysis of features observed from radar images of the surface (e.g., Masursky et al. 1980, Campbell et al. 1983, 1984, Barsukov et al. 1986, Basilevsky et al. 1986). In a previous study (Zuber 1987), we developed a theoretical approach for relating the geometries of tectonic surface features on Venus to the structure of the lithosphere. We demonstrated how the widths and spacings (wavelengths) of tectonic features could be used in combination with theoretical models of compressional and extensional deformation of the lithosphere to yield constraints on Venus’ crustal thickness and near-surface thermal gradient.

The objective of the present study is to analyze the relationship between wavelength and regional isostatic surface elevation, two parameters that can be observed with existing data. We accomplish this by first relating elevation and wavelength to crustal thickness and thermal gradient, and then relating elevation and wavelength to each other. We show that in areas of Venus where the upper mantle is stronger than the upper crust, long wavelengths ($\lambda \approx 30$ km) of deformation, which develop mainly in response to unstable deformation of the strong upper mantle, are not simply related to elevation due to the effect of crustal thickness on the depth distribution of lithosphere strength. However, short wavelengths ($\lambda \approx 30$ km) of deformation, which develop primarily due to unstable deformation of the upper crust, may either increase or remain constant with increasing isostatic elevation if the elevation increase is due to increasing crustal thickness. In addition, short-wavelength spacings, for either a strong or weak upper mantle, should decrease with increasing elevation if the elevation increase is due to increasing surface thermal gradient. These relationships demonstrate that correlation of wavelength with elevation for short wavelengths of tectonic deformation on Venus may yield useful information on spatial variations in lithosphere structure by providing a basis for distinguishing elevation changes due to differences in crustal thickness and surface thermal gradient.

TECTONIC FEATURES, ISOSTATIC ELEVATION, AND LITHOSPHERE STRUCTURE

Tectonic features in many areas of Venus (e.g., Fig. 1) consist of subparallel lineations (Campbell et al. 1983, 1984, Barsukov et al. 1986, Basilevsky et al. 1986) characterized by regular widths or spacings (Campbell et al. 1983, Solomon and Head 1984, Zuber 1987). The geometries of these features provide information on the state of stress and rheological stratification of the lithosphere. Tectonic features characterized by a single spacing or dominant wavelength can be explained by horizontal extension or compression of a lithosphere with a strong surface layer underlain by a weaker continuum (Solomon and Head 1984, Zuber and Parmentier 1986, Zuber 1987, Banerdt and Golombek 1988), while features characterized by more than one dominant wavelength require a lithosphere that consists of a strong surface layer underlain successively by a weaker layer and a layer stronger than or of nearly comparable strength to the surface layer (Zuber et al. 1986, Zuber 1987, Banerdt and Golombek 1988).

The dominant wavelengths of tectonic features are sensitive to the thickness of the crust and the thermal structure of the lithosphere. Increasing the thermal gradient reduces the depths of brittle–ductile transitions in the crust and mantle, and decreases the strengths and thicknesses of strong regions within the lithosphere. Increasing the crustal thickness increases the thickness of
Fig. 1. Venera 15/16 mosaic of an area in the northern hemisphere of Venus that contains ridge belts, which are tectonic features that exhibit two parallel length scales. The longer length scale corresponds to the wavelength of the belts measured normal to their trend, and the shorter length scale is defined by the spacings of ridges and grooves within individual belts. The ridge belts are located in the upper half of the image and trend approximately north–south. The quasicircular feature near the center of the image has a diameter of about 175 km.
the weak lower crust and correspondingly decreases the thickness of a strong upper mantle region. Variations in thermal gradient or crustal thickness can explain differences in the wavelengths of tectonic features observed in different areas, while crustal thickness variations can provide a simple explanation for the existence of single or multiple wavelengths of deformation. Areas that exhibit multiple wavelengths can be explained if the lithosphere consists of a relatively thin crust that is strong near the surface and weaker at depth, underlain by an upper mantle that is much stronger than the lower crust (Zuber 1987, Banerdt and Golombek 1988). Such a structure is also suggested by models of viscous relaxation of Venusian surface topography constrained by the depths of large-impact craters (Grimm and Solomon 1988). Areas that exhibit a single tectonic wavelength can be explained by a lithosphere with a thick crust that does not contain a region of upper mantle strength.

Regional isostatic elevation, characterized by topography that is not supported by either the strength of the lithosphere or a dynamic mechanism such as mantle convection, is also related to thermal structure and crustal thickness. A high thermal gradient and/or a thick crust correspond to a high isostatic surface elevation, while a low thermal gradient and/or a thin crust correspond to a low surface elevation. In areas of relatively shallow compensation, i.e., in the absence of a dynamic component of support, isostatic topography and the dominant wavelengths of tectonic features should be correlated. The purpose of this study is to quantify this relationship and explore the implications for Venus’ lithosphere structure.

**APPROACH**

We proceed by establishing, in turn, the relationships of wavelength and elevation with crustal thickness and surface thermal gradient. We then relate wavelength and elevation to each other. Previous studies that have treated the rheology of the Venus lithosphere have arbitrarily assumed simple linear thermal gradients (Zuber, 1987, Banerdt and Golombek 1988, Grimm and Solomon 1988). We prefer a more general and arguably more realistic representation of the thermal structure in which the distribution of temperature \( T \) with depth \( z \) in the lithosphere is estimated from

\[
T(z) = T_s + (T_m - T_s) \text{erf}(z/\delta) \tag{1}
\]

where \( T_s \) is the surface temperature, \( T_m \) is the mantle temperature, and \( \delta \) is the characteristic depth of the temperature structure. Equation (1) is a convenient mathematical description for a lithosphere in which temperature increases with depth near the surface and approaches a limiting value at depth. The role of the surface thermal gradient \( (dT/dz)_0 \)

\[
\frac{dT}{dz_0} = \frac{2(T_m - T_s)}{\delta \pi^{1/2}} \tag{2}
\]

can be examined by simply varying \( \delta \).

Thermal gradients determined from this general representation of the thermal structure can be compared to those predicted by more specific thermal models. For example, consider a lithosphere that acts as a conducting, steady-state thermal boundary layer and that contains a crust with radiogenic heat-producing elements. Further, assume that the lithosphere is underlain by a convecting asthenosphere that supplies a constant heat flux \( q_m \) to the base of the lithosphere. The temperature distribution is described mathematically by

\[
T(z) = T_s + \left[ \frac{q_m}{k_m} + \frac{H_c c}{k_c} \right] z - \frac{H_c}{2k_c} z^2 \\
0 < z < c \tag{3a}
\]

\[
T(z) = T_s + \frac{q_m}{k_m} z + \frac{H_c c^2}{2k_c} \\
c < z < L \tag{3b}
\]

where \( H_c \) is the rate of heat production for basalt, \( k_c \) and \( k_m \) are the thermal conductivities of the crust (basalt) and mantle (olivine), \( c \) is the thickness of the crust, and \( L \)
TABLE I

MODEL PARAMETERS

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Unit</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>$T_s$</td>
<td>700</td>
<td>K</td>
<td></td>
</tr>
<tr>
<td>$T_m$</td>
<td>1330</td>
<td>K</td>
<td></td>
</tr>
<tr>
<td>$\rho_0$</td>
<td>0</td>
<td>kg m$^{-3}$</td>
<td></td>
</tr>
<tr>
<td>$\rho_c$</td>
<td>2900</td>
<td>kg m$^{-3}$</td>
<td></td>
</tr>
<tr>
<td>$\rho_m$</td>
<td>3300</td>
<td>kg m$^{-3}$</td>
<td></td>
</tr>
<tr>
<td>$q_m$</td>
<td>74</td>
<td>mW m$^{-2}$</td>
<td></td>
</tr>
<tr>
<td>$H_c$</td>
<td>$1.0 \times 10^{-7}$</td>
<td>W m$^{-3}$</td>
<td></td>
</tr>
<tr>
<td>$\alpha$</td>
<td>$3.0 \times 10^{-5}$</td>
<td>K$^{-1}$</td>
<td></td>
</tr>
<tr>
<td>$k_c$</td>
<td>2</td>
<td>W m$^{-1}$ K$^{-1}$</td>
<td></td>
</tr>
<tr>
<td>$k_m$</td>
<td>4</td>
<td>W m$^{-1}$ K$^{-1}$</td>
<td></td>
</tr>
<tr>
<td>$g$</td>
<td>8.87</td>
<td>m sec$^{-2}$</td>
<td></td>
</tr>
</tbody>
</table>

$^a$ Solomon and Head (1982).

$^b$ Turcotte and Schubert (1982).

is the thickness of the lithosphere defined, as in many terrestrial studies, by the $0.9T_m$ isotherm. For values thought to be appropriate for Venus (see Table I), $dT/dz_0 = 18^\circ$K km$^{-1}$.

Isostatic surface elevation ($d$) is calculated on the basis of the thermal and compositional structure of the lithosphere, which depends on the temperature distributions given above and the thickness of the crust. The elevation is also dependent on the parameters $\rho_c$ and $\rho_m$, the column densities of the crust and mantle, and $\alpha$, the volume coefficient of thermal expansion. Parameter values which we have adopted are given in Table I.

We determine the relationships between the wavelengths of tectonic features and crustal thickness and thermal gradient following the study of Zuber (1987). This study demonstrated that, for a range of physically plausible rheologically stratified lithosphere structures, uniform horizontal compression or extension results in the growth of small-amplitude instabilities that amplify with time and result in deformation with a dominant wavelength. Model parameters in the present study are specified on the basis of strength envelopes for the Venus lithosphere calculated for a range of $c$ and $dT/dz_0$. The lithosphere is assumed to consist of a diabase crust that overlies an olivine mantle. Brittle deformation of the crust and upper mantle is treated by employing a perfectly plastic rheology. Ductile deformation of the crust is treated using Caristan's (1982) flow law for Frederick diabase, which is compositionally similar to the Venus surface at the Venera 13 and 14 landing sites (Surkov et al. 1984). Ductile deformation of the mantle is treated using a flow law for dry, polycrystalline olivine (Goetze 1978).

For all models the strong upper crust is assumed to have a brittle rheology and a thickness corresponding to the depth of the brittle–ductile transition. For the strong upper mantle, two different cases are considered. In the first, the mantle is assumed to deform in a brittle manner and is modeled by a plastic layer with a thickness that depends on the depth of the mantle brittle–ductile transition. In the second, deformation of the strong upper mantle is assumed to be governed primarily by ductile dislocation creep. The thickness of a strong, ductile mantle layer depends on the depth at which strength decreases by a factor of $1/e$ compared to its value at the mantle brittle–ductile transition. However, models which assume that the ductile layer thickness equals the sum of the $e$-folding thickness and the mantle brittle zone, or this sum divided by 2, do not yield markedly different results.

The dimensionless parameter $S_1 = [(\rho_1 - \rho_0)gh_1/\tau_1]$ characterizes the ratio of the buoyancy forces that arise due to layer thickness variations to the characteristic strength of the crustal layer. $S_1$ depends on the density ($\rho_1$), thickness ($h_1$), and strength ($\tau_1$) of the surface (strong upper crustal) layer, the density of the continuum that overlies the crust ($\rho_0$), and the acceleration of gravity ($g$) on Venus. Results are determined for $S_1 = 0$, which corresponds to the limit of a strong surface layer in which buoyancy forces do not play an important role in the deformation, and for $S_1 = 10$, which corresponds to the limit of a
Fig. 2. Example plot of dimensionless growth rate \( q \) as a function of wavenumber \( (k' = 2\pi h_1/\lambda, \text{bottom axis}) \) and wavelength \( (\lambda/h_1, \text{top axis}) \) for a strength-stratified model lithosphere in extension. (The parameter \( h_1 \) represents the thickness of the strong upper crust.) The wavenumbers or wavelengths at which extensional instabilities grow fastest (i.e., the dominant wavelengths) are represented by maxima in \( q \). The magnitude of \( q_{\text{max}} \) associated with a given dominant wavelength indicates the rate at which deformation will develop at that wavelength. In this example, two wavelengths of extensional deformation are likely to develop because the growth rates associated with the dominant wavelengths have nearly identical magnitudes.

weak surface layer in which buoyancy forces significantly affect the characteristics of deformation.

Dominant wavelengths due to unstable compression or extension of a rheologically stratified medium are defined by maxima in a growth rate rate spectrum (e.g., Fletcher 1974). The magnitude of the growth rate, \( q \), associated with a given wavelength indicates the exponential rate at which deformation will develop. In previous studies that considered the development of multiple dominant wavelengths of tectonic features (Ricard and Froidevaux 1986, Zuber et al. 1986, Zuber 1987), no consideration was given to the relative magnitudes of the maximum growth rates. These studies addressed only whether the amplitude of an individual growth rate was large enough such that an instability would amplify with time. However, if multiple dominant wavelengths are to develop, then the corresponding dominant growth rates should be of comparable magnitudes, as illustrated in Fig. 2. We therefore modify our previous approach for determining the number of wavelengths that will occur by imposing a growth rate similarity requirement for the development of multiple wavelengths. Because it is difficult to assess how "similar" the growth rates must be for multiple wavelengths of deformation to grow, we show results arbitrarily assuming that dominant growth rates must be within a factor of 3 of each other. We then discuss limiting cases and demonstrate that solutions are largely insensitive to the exact numerical value of the similarity requirement.

RESULTS

Relationships between \( \lambda, c, \) and \( dT/dz_0 \)

Model constraints are provided by the widths and spacings of tectonic features, which are listed in Table II. Of the multiple wavelength features, the rift zones are interpreted as extensional (Pettengill et al. 1979, Masursky et al. 1980, Campbell et al. 1984), while the ridge belts have been interpreted as both compressional (Barsukov et al. 1986, Zuber 1986, Frank and Head 1988) and extensional (Sukhanov 1987, Sukhanov and Pronin 1988). The banded terrain in Akna, Freyja, and Maxwell Montes, which displays a single short wavelength of deformation, is interpreted as compressional (Campbell et al. 1983, Vorder Bruegge et al. 1988).

Figures 3–5 summarize the limits on crustal thickness \( (c) \) and surface thermal gradient \( (dT/dz_0) \). The combined black and hatched boxes represent solutions obtained using the method from our previous study (Zuber 1987). The black boxes alone represent solutions in which dominant growth rates must be within a factor of 3. Figure 3 shows the results for rift zones, assuming that these features formed due to lithospheric extension. For a strong surface layer, successful models (black boxes) for the assumed growth rate similarity require-
TABLE II
LENGTH SCALES OF TECTONIC FEATURES ON VENUS

<table>
<thead>
<tr>
<th>Tectonic feature</th>
<th>Long-wavelength component</th>
<th>Long-wavelength width or spacing (km)</th>
<th>Short-wavelength component</th>
<th>Short-wavelength width or spacing (km)</th>
<th>Reference*</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rift zones</td>
<td>Rift zone width</td>
<td>70–250</td>
<td>Ridges and grooves</td>
<td>10–20</td>
<td>1, 2</td>
</tr>
<tr>
<td>Ridge belts</td>
<td>Belt spacings</td>
<td>150–350</td>
<td>Ridges and grooves</td>
<td>10–20</td>
<td>3, 4</td>
</tr>
<tr>
<td>Banded terrain</td>
<td></td>
<td></td>
<td>Bands/ridges</td>
<td>8–20</td>
<td>5, 3</td>
</tr>
</tbody>
</table>

* 1 = Campbell et al. (1984); 2 = Schaber (1982); 3 = Basilevsky et al. (1986); 4 = Zuber (1986); 5 = Campbell et al. (1983)

![Diagram](image)

**FIG. 3.** Ranges of thermal gradient \((dT/dz_o)\) and crustal thickness \((c)\) in areas of the Venus surface that contain rift zones, which were assumed to form due to lithospheric extension. Boxes indicate parameter ranges for which deformation occurs with multiple wavelengths, while areas without boxes indicate deformation with a single wavelength. Results are shown for (a) strong \((S_t = 0)\) and (b) weak \((S_t = 10)\) lithosphere models with brittle \((n = \infty)\) and viscous \((n \approx 3)\) strong mantle layers. Unfilled boxes correspond to combinations of \(c\) and \((dT/dz_o)\) for which theoretically predicted values of multiple dominant wavelengths are inconsistent with the ranges of wavelengths listed in Table II. Hatched boxes correspond to model results in which predicted multiple wavelengths are consistent with observations but dominant growth rates of short- and long-wavelength instabilities differ by more than a factor of 3. Black boxes correspond to models in which the predicted multiple dominant wavelengths are consistent with observations and the dominant growth rates are within a factor of 3.
ment are obtained for brittle and viscous mantle layers for $c < 15$ km and $dT/dz_0 \leq 10^\circ$K km$^{-1}$. If the surface layer is weak, then only a brittle mantle layer can produce multiple wavelengths of deformation, and limits of $c < 15$ km and $dT/dz_0 \leq 25^\circ$K km$^{-1}$ are obtained. Note that $c$ and $dT/dz_0$ trade off in a manner such that for a given wavelength or range of wavelengths, lower thermal gradients are compatible with larger crustal thicknesses and vice versa.

The results for ridge belts are shown in Figs. 4 and 5. Because there is debate concerning whether these features are compressional or extensional, results are shown for both cases. The compressional models yield successful solutions only for a viscous mantle layer and strong surface layer, and indicate $c < 20$ km and $dT/dz_0 \leq 10^\circ$K km$^{-1}$. The extensional models yield successful solutions only if the surface layer is weak and the mantle layer deforms primarily in a brittle manner. As for the compressional case, values of $c \leq 20$ km and $dT/dz_0 \leq 10^\circ$K km$^{-1}$ are implied.

If the growth rate similarity requirement is relaxed, then larger maximum values of the crustal thickness and thermal gradient are possible. For example, in the limiting case in which the requirement is removed completely, allowable ranges of $c$ and $dT/dz_0$ are defined by the combined hatched and black boxes. Imposition of a more stringent growth rate similarity requirement than assumed in Figs. 3–5 shifts the maximum allowable solutions for $c$ and $dT/dz_0$ to lower values.

The imposition of the growth rate similarity requirement does not significantly modify the results of our previous study (Zuber 1987). The range of allowable surface thermal gradients for rift zones ($dT/dz_0 \leq 25^\circ$K km$^{-1}$) is compatible with mean lithospheric thermal gradients determined from models.
FIG. 4. Ranges of $dT/dz_0$ and $c$ in areas of the Venus surface that contain ridge belts, which were assumed to form due to lithospheric compression. Results are shown for (a) strong ($S_i = 0$) and (b) weak ($S_i = 10$) lithosphere models with brittle ($n = \infty$) and viscous ($n = 3$) strong mantle layers. Box shadings are defined in the legend to Fig. 3.
FIG. 5. Ranges of $dT/dz_o$ and $c$ in areas of the Venus surface that contain ridge belts, which were assumed to form due to lithospheric extension. Results are shown for (a) strong ($S_l = 0$) and (b) weak ($S_l = 10$) lithosphere models with brittle ($n = \infty$) and viscous ($n \approx 3$) strong mantle layers. Box shadings are defined in the legend to Fig. 3.
of hot spot-dominated heat loss, 10°K km⁻¹ (Morgan and Phillips 1983), global scaling arguments, 16–24°K km⁻¹ (Soloman and Head 1982, Grimm and Solomon 1988), and steady-state conduction, 18°K km⁻¹ [Eq. (3)], as well as with measured values in old terrestrial ocean basins and continental shields, 10°K km⁻¹ (Sclater et al. 1980). Model results for the ridge belts are consistent with the lower end of the above range of thermal gradients (dT/dz₀ ≤ 10°K km⁻¹).

The upper limit of the allowable crustal thickness on Venus (=20 km if growth rates must be within a factor of 3) is in agreement with, though somewhat lower than, our earlier estimate which did not incorporate a growth rate similarity requirement, 30 km (Zuber 1987). It also agrees well with other estimates based on the geometries of tectonic features, 5–15 km for 10 < dT/dz < 15°K km⁻¹ (Banerdt and Golombek 1988), the depths of impact craters, 10–20 km for dT/dz > 10°K km⁻¹ (Grimm and Solomon 1988), and models for gravity-driven crustal decollements, <34 km for dT/dz > 10°K km⁻¹ (Smrekar and Phillips 1988). However, this estimate is much lower than suggested by compensation depths of long-wavelength topography (Phillips et al. 1981, Phillips and Malin 1984), parameterized convection models (Solomatov et al. 1987), or the depth of the basalt/eclogite phase transition on Venus (Anderson 1980, Kaula 1988), all of which yield c on the order of 100 km. However, the latter estimates may not accurately reflect the thickness of the crust if dynamic compensation is an important means of long-wavelength topographic support on Venus, if boundary or initial conditions describing Venus' thermal evolution were different than assumed by Solomatov et al. (1987), or if crustal differentiation occurred at depths shallower than the basalt/eclogite transition.

The limits on dT/dz₀ and c indicated in Figs. 3–5 are valid only for areas in which tectonic features with multiple wavelengths are found. In other areas, such as the banded terrain where only a single, short wavelength of deformation (cf. Table II) is observed, a lithosphere with a crust thick enough such that the mantle lacks a region of strength is required. For these areas limits of c ≥ 20 km are implied if dominant growth rates are constrained to be within a factor of 3 or c ≈ 30 km if no growth rate similarity requirement is imposed.

The above-mentioned constraints on lithosphere structure are also valid only for areas in which multiple length scales of deformation developed contemporaneously. On the basis of superposition relationships in the Venera 15 and 16 radar images, Sukhanov (1987) and Sukhanov and Pronin (1988) have suggested that structures within the ridge belts locally exhibit an age progression. To interpret this observation in the context of our analysis it is instructive to emphasize that our linearized models describe the first increment of deformation, when the dominant length scales developed. The models are thus consistent with a scenario in which progressive episodes of finite deformation developed with length scales controlled by an early phase of unstable deformation. We also note that we are aware of no other mechanisms that can simply explain the development of multiple parallel length scales of deformation on Venus within the context of current information of Venus’ shallow internal compositional and rheological structure.

Relationships between d, c, and dT/dz₀

Relationships between isostatic elevation (d), crustal thickness, and surface thermal gradient are shown in Fig. 6, which was derived for an isostatic balance with the temperature distribution in Eq. (1). Note that a thicker crust and higher thermal gradient both correspond to a higher elevation. However, for ranges of c and dT/dz₀ that may be plausible for Venus, significant differences in isostatic elevation can most easily be explained by crustal thickness variations. For example, a 10°K km⁻¹ change in dT/dz₀, which would imply a significant change in thermal structure on the regional
scale, corresponds to less than a half-kilometer change in elevation, while a presumably modest 10-km change in \( c \) corresponds to about a 1-km elevation difference.

**Relationships between \( \lambda, d, c, \) and \( dT/dz_0 \)**

Figures 7 and 8 show relationships between elevation and dominant wavelength \( (\lambda_d) \) for extensional and compressional models. To distinguish between the effects of elevation changes due to thermal and compositional effects, we show results for two surface thermal gradients, 10 and 20°K km\(^{-1}\). Any change in elevation for a given value of \( dT/dz_0 \) is a consequence of variation in crustal thickness. We first consider results for short-wavelength \( (\lambda \approx 30 \text{ km}) \) deformation for the strong mantle case.

Short-wavelength deformation is considered in Figs. 7 and 8. For a given crustal thickness, increasing \( dT/dz_0 \) decreases the dominant wavelength for short-wavelength extensional and compressional tectonic features. This is a consequence of reducing the thickness of the strong upper crust. The dominant wavelength, which is directly proportional to the thickness of the strong layer, therefore also decreases.

For constant \( dT/dz_0 \), both short-wavelength compressional and extensional tectonic features show a clear trend of increasing wavelength with elevation. This holds for both strong \( (S_1 = 0) \) and weak \( (S_1 = 10) \) surface layers as well as for both viscous and brittle mantle layers. (The pattern may be modified for some cases at low isostatic elevations in which multiple wavelengths of short-wavelength deformation develop). Wavelength increases with elevation for a range of weak lower crustal thicknesses because the short-wavelength of instability, though controlled primarily by the thickness of the strong upper crust and the strength contrast between the upper and lower crust, is also influenced by the mechanical properties of the strong upper mantle. The mantle resists deformation at these short wavelengths, which causes the dominant wavelength to progressively decrease as the depth to the crust–mantle boundary decreases (cf. Zuber and Aist 1990).

Figures 7 and 8 show that at isostatic elevation changes of greater than about 1 km, the dominant wavelengths of the short-wavelength features in most cases approach a limiting value. Over the elevation range where this occurs the mechanical properties of the strong upper mantle do not markedly influence the short-wavelength deformation, and for a given value of \( dT/dz_0 \) wavelength remains nearly constant with elevation. The lithosphere in this case consists of a strong upper crust, a thick weak lower crust, and an upper mantle with strength less than or equal to the strength of the upper crust. Extension or compression then results in a single wavelength of deformation controlled by the thickness and stress exponent of the strong upper crust and by the strength contrast between the upper and lower crust. For a lithosphere that deforms with a single dominant wavelength, the wavelength should remain constant with elevation changes due to changing crustal thickness.

Table III summarizes the relationships between \( dT/dz_0, c, d \) and \( \lambda \) for short-wavelength features that occur in areas with
strong and weak mantle layers. These relationships, which are valid for both extension and compression, provide a basis for discerning spatial variations in Venus' lithosphere structure. Specifically, in areas of shallowly compensated isostatic topography where reliable measurements of elevation exist and short-wavelength spacings can be measured with relative certainty, it should be possible to distinguish whether isostatic elevation differences are a consequence of variations in crustal thickness or thermal gradient. The correlation cannot be applied in highland regions such as Beta Regio and some parts of Aphrodite Terra which are deeply and probably dynamically compensated.

Information on the nature of compensa-
Fig. 8. Variation of dominant wavelength ($\lambda_d$) with isostatic elevation ($d$) for the Venus lithosphere in compression assuming (a) strong ($S_l = 0$) and (b) weak ($S_l = 10$) lithosphere models with brittle ($n = \infty$) and viscous ($n = 3$) strong mantle layers.

The relationship between $\lambda$ and $d$ due to changes in $c$ for long-wavelength features is more complicated than for short-wavelength ones. At these long wavelengths, the structure of the entire lithosphere influences the pattern of deformation, and for a given value of $dT/dz_0$ wavelength may either increase, remain constant, or decrease with elevation (see Figs. 7 and 8). However, in any case, the long wavelength of deformation vanishes at high isostatic ele-
TABLE III

<table>
<thead>
<tr>
<th>Increase (dT/dz_0)</th>
<th>Increase (c)</th>
</tr>
</thead>
<tbody>
<tr>
<td>When the upper mantle is stronger than the upper crust</td>
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<td>(d) Increases</td>
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<tr>
<td>(d) Increases</td>
<td>Increases</td>
</tr>
<tr>
<td>(\lambda) Decreases</td>
<td>No change</td>
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</tbody>
</table>

vation (i.e., large crustal thickness) because the mantle does not contain a region of strength.

Lithosphere Stresses and the Origin of Ridge Belts

Due to differences in interpretation of observational data, controversy has arisen regarding whether the ridge belts are compressional or extensional in nature. Possible insight into the origin of ridge belts can be obtained by estimating the predicted levels of stress in the upper crustal layer for extensional and compressional instability models. Upper and lower limits of the layer strength can be made by observing the values of \(S_1\) at which limiting behavior sets in for a strong (small \(S_1\)) and weak (large \(S_1\)) upper crustal layer (cf. Fig. 7 in Zuber 1987). The results indicate that if ridge belts are compressional, then the upper crust should have an average strength on the order of 150 MPa or more, while if they are extensional, then strength should be on the order of 15 MPa or less. If the experimental data on rock fracture and flow assumed in this study (i.e., Byerlee 1968, Caristan 1982) reasonably approximate the rheological properties of the Venus crust at the time of ridge belt formation, then these values are most compatible with a compressional origin for the ridge belts. However, because of uncertainties in the rheological structure of the Venus lithosphere, any conclusions based on absolute strengths should be interpreted with appropriate caution. It should also be noted that a compressional origin for the ridge belts as a whole would not rule out the possibility that some ridge belts formed as a consequence of localized extension.

Implications for the Nature of Viscous Relaxation of Topography

Because of the high surface temperature and predicted low erosion rates on Venus (Garvin et al. 1984, Ivanov et al. 1986), viscous relaxation of topography may be the dominant mechanism for the reduction of surface relief (Weertman 1979, Solomon et al. 1982, Bindshadler and Parmentier 1987, Grimm and Solomon 1988). An increase in crustal thickness and/or and thermal gradient decreases lithosphere viscosity, resulting in a more rapid rate of relaxation. Therefore, surface features on Venus at higher isostatic elevations would be expected to relax faster than those at lower elevations. In areas where deformation has been nearly contemporaneous, surface features at higher elevations should appear more highly degraded than those at lower elevations.

Observed Correlation of Wavelength with Elevation

Figure 9, reproduced from Ivanov (1988), summarizes an observed correlation of wavelength with elevation for short-wavelength ridges in several areas of Venusian tessera terrain. It is not certain whether the topography of these areas is due to isostatic changes in crustal thickness or thermal structure, as assumed in our models, because gravity data have not been analyzed. However, no significant gravity anomaly is observed in Tellus Regio (Sjogren et al. 1983), another area of tessera. If this area is representative of other tessera, then the topography of the tessera may be isostatically compensated at shallow depths rather than dynamically supported.
We limit discussion of observations in Fig. 9 to Laima Tessera and the other smaller tessera; these areas show an increase of ridge spacing with elevation. The single measurement for Fortuna Tessera is difficult to interpret because Ivanov averaged observations in this area for a variety of different tectonic subunits (defined by Vorder Bruegge and Head 1988) distributed over a range of elevations. Assuming isostatic compensation due to variations in crustal thickness or thermal structure, the observed relationship between ridge spacing and elevation interpreted in the context of our analysis (Table III) indicates a thicker crust in areas with the larger spacing. The observations also imply the presence of a strong upper mantle beneath Laima and the other tessera regions.

Quantifying and understanding the relationship between ridge spacing and elevation in complexly deformed Fortuna Tessera will require more detailed analysis. Vorder Bruegge and Head (1988) have established a correlation between surface elevation and the intensity of tectonic deformation in this area, and interpret the results in terms of crustal thickening due to large-scale compression. If crustal thickness varies, then the spacings of tectonic features within the various geomorphic-tectonic units in Fortuna should be related to elevation as predicted in Table III, unless the thermomechanical structure and stress history of this region are more complicated than assumed in our simple models. For example, if the Fortuna Tessera/Maxwell Montes area has undergone significant volcanic activity, then there could be subsurface density variations caused by factors other than those treated in this study that could contribute to the high topography.

CONCLUSIONS

To better understand the lithosphere structure of Venus, we have developed models that relate surface thermal gradient and crustal thickness to regional isostatic elevation and the wavelengths of tectonic surface features. We have added an additional constraint to our previous approach for determining the relationship between multiple wavelengths of deformation and lithosphere structure (Zuber et al. 1986, Zuber 1987) by requiring that for multiple wavelengths to develop, the growth rates of instabilities that determine the wavelengths must be "similar" in magnitude. This requirement constrains the maximum values of crustal thickness and thermal gradient in areas with multiple wavelengths to be less than or equal to those determined in our previous study of tectonic deformation on Venus, \( c \leq 30 \text{ km} \) and \( dT/dz_0 \leq 25^\circ \text{K km}^{-1} \) (Zuber 1987). These results are sensitive to the applicability of experimentally derived flow data to the compositions and strain rates that characterize Venus.

In areas of Venus where the upper mantle is stronger than the upper crust, the spacings of the short-wavelength (\( \lambda \approx 30 \text{ km} \)) features, formed primarily due to unstable extension or compression of the strong upper crust, should increase with increasing isostatic elevation if the elevation increase is due to increasing crustal thick-
ness. However, this wavelength should decrease with increasing elevation if the elevation increase is due to increasing thermal gradient. Long-wavelength (λ ≥ 30 km) features, formed mainly as a result of unstable deformation of the strong upper mantle, should exhibit more complicated relationships between λ and d primarily because of the influence of crustal thickness on the depth distribution of lithospheric strength. As a consequence, an analysis of the relationships between λ and d for observed long-wavelength features would probably not be worthwhile.

In areas of Venus where the upper mantle is weaker than the upper crust, the spacings of short-wavelength features should be independent of elevation if the elevation change is due to variations in crustal thickness, but should decrease with increasing elevation if the elevation increase is due to increasing thermal gradient.

The results of this study demonstrate the systematic correlation of short-wavelength (λ ≤ 30 km) spacings with elevation, for tectonic features that develop in areas with either a strong or weak upper mantle, could provide useful information on spatial variations in Venus' surface thermal gradient and crustal thickness. Such analyses may, for at least some areas where wavelength and elevation can be measured with relative certainty, provide a valuable means of discerning which of these effects is primarily responsible for variations in regional scale isostatic topography. For example, a preliminary analysis that correlated spacing with elevation in several areas of tessera (Ivanov 1988) has yielded evidence for changes in crustal thickness. Currently available Arecibo and Venera 15 and 16 data are of suitable coverage and resolution for regional studies, whereas data from the Magellan mission will be well suited for global correlations.

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REFERENCES


Garvin, J. B., J. W. Head, M. T. Zuber, and P. Helfenstein 1984. Venus: The nature of the sur-


ZUBER, M. T. 1987. Constraints on the lithospheric structure of Venus from mechanical models and tec-

